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Observations on Drought and Vegetation:
A Global Approach and Local Study of the Congo Basin

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Observations on Drought and Vegetation: A Global Approach and Local Study of the Congo Basin

by

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To all who helped me even outside of science:

My parents, two brothers, friends, and two dogs.

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Observations on Drought and Vegetation: A Global Approach and Local Study of the Congo Basin

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Current research suggests that climate change and continued anthropogenic forcing of the climate system may increase the occurrence of extreme events such as droughts in the near future. Prolonged droughts are potentially destructive because of widespread impacts and difficulty in prediction. In addition, droughts interact in complex ways with vegetation systems, inducing abrupt rather than gradual changes in vegetation. Using satellite and station rainfall data, this research seeks to identify patterns in drought and vegetation change on a global scale and in the tropics (15°N to 35°S), where precipitation is the predominant determinant of climate. It is observed that long drought occurs in regions of variability in vegetation. Additionally, long drought is observed in regions on the fringe between dominant vegetation types of desert, grassland, and forest. These results suggest that prolonged drought may be

a significant contributor to abrupt vegetation changes. The other part of this study is a regional investigation of vegetation change and drought in the Congo Basin. The correlation between drought and forest loss in the Congo is not clear. Preliminary studies of the atmospheric mechanisms governing drought in the Congo are presented. The results of this work as a whole will provide a better understanding of drought patterns and vegetation responses to drought, ultimately to provide better adaptation to a more extreme future climate.

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Chapter 1

Introduction

In light of recent changes in climate and further anthropogenic forcing of the climate system, it has been suggested that extreme weather and climate events may increase in occurrence in the near future [e.g., 20, 3]. Among extreme events, prolonged droughts are potentially destructive due to large impacts on widespread areas and the difficulties involved in predicting such events [e.g., 3, 18]. For example, the 2012 Great Plains drought in the United States cost over $10 billion in agricultural losses and other damages, indicating the scale of potential drought impacts [18].

Extreme droughts are also significant in shaping ecosystems due to feedback between rainfall and vegetation. It has been suggested that vegetation responds abruptly rather than linearly to changes in rainfall [e.g., 40]. In this theory, a drought (abnormally low rainfall) exceeding a certain threshold (dependent on the system) triggers a sudden change in a vegetation state. For example, a dry season lasting one month longer than usual may trigger the death of a forest and its replacement by grassland. By analogy, greater than normal rainfall above a certain threshold is required to return the system to its original state (e.g., return grassland to a forested state). As a result, drought may permanently alter a vegetation ecosystem if the wet threshold is not met to restore the previous ecosystem.

By extension, Hirota et al. (2011) found observational evidence that tropical vegetation tends to exhibit three distinct stable states of desert, grassland, and forest [17]. Distributions of tree cover plotted against climatological annual rainfall in the tropics (15°N – 35°S) shows these distinct stable states universally in all the continents (South America, Africa, and Australia). This observation supports a double hysteresis rainfall–vegetation feedback model in which the stable states are basins of attraction and the two regions in between are unstable[1]. This discussion of abrupt vegetation responses to rainfall warrants further study and so we will return to it in a later section in this chapter.

While the physical mechanisms behind the aforementioned rainfall–vegetation feedback system are not clear, some theories have been suggested. For example, Charney (1975) theorized that changes in albedo between forested regions and barren land have a significant impact on atmospheric circulation [7]. Transpiration of plants, particularly trees with deep and complex root systems, may also contribute to the feedback between vegetation and rainfall by increasing the moisture content of the atmosphere [21].

The importance of understanding the relationship between vegetation cover, land changes, rainfall, and drought is significant and motivates this

[1]These instabilities may also be referred to as bifurcation points.

particular study. In this work, we use observational data to try to establish links between the various theories of abrupt vegetation changes. In particular, we are interested in drought as a governing mechanism for vegetation changes on a global scale. We also attempt to demonstrate the global characteristics of drought in a regional study in the Congo basin. If the importance of drought can be shown in determining abrupt vegetation changes on a global scale, this result could have great importance in modeling vegetation changes in a future, warmer climate.

The remainder of this chapter is dedicated to a background review of current knowledge of vegetation feedback systems. Theoretical treatments of the feedback systems are discussed along with observational evidence. We also present the characteristics of the Congo basin that are of particular interest in focusing on the region.

1.1 Rainfall–vegetation feedback systems: Theory

Until recently, it was assumed (notably in climate models) that land cover (vegetation) changed linearly in response to precipitation anomalies. Thus, if a region experienced an abnormal increase in precipitation, more trees might begin to grow, and once precipitation returned to normal, these trees would no longer be sustained. Conversely, an abnormal decrease in precipitation would cause trees and other vegetation to die back, but these plants would return upon resumption of normal rainfall. Empirical evidence, however, has disproved this linearity and rather suggested that the resumption of normal rainfall conditions may not be a sufficient condition for restoration of the previous vegetation state [40]. This condition is the basis for nonlinearity: rainfall anomalies exceeding a threshold shift the vegetation state from one stable state to another. This threshold, however, is not the same as the threshold for returning to the original stable state. The result is shown schematically in Figure 1.1.

As indicated previously, there are many physical factors that determine this effect. The first, suggested by Charney (1975) [7], involves the difference in albedo (the fraction of incoming solar radiation that is reflected) between vegetated and barren land. Barren land has a higher albedo than vegetated land, and thus does not heat as much. The result is a relatively cooler lower atmosphere due to less ground absorption of radiative heat. The cooler lower atmosphere is more stable and induces a general sinking motion in a vertical atmospheric column. This state inhibits convection, the main process by which rainfall develops in the tropics. Thus, a lack of vegetation reinforces the lack of rainfall and produces a self-sustaining desert environment. For a vegetated environment, a warmer lower atmosphere not only induces more vertical instability (rising motion) and thus convective precipitation but also increases the mechanisms by which monsoons occur[2]. Monsoons result from a temperature gradient between ocean and land as the land warms faster than the ocean during the day. As a result, cool, moist air is advected (transported

[2]See section 4.2 for more information on the effect of monsoons on vegetation and the positive feedback system involved.

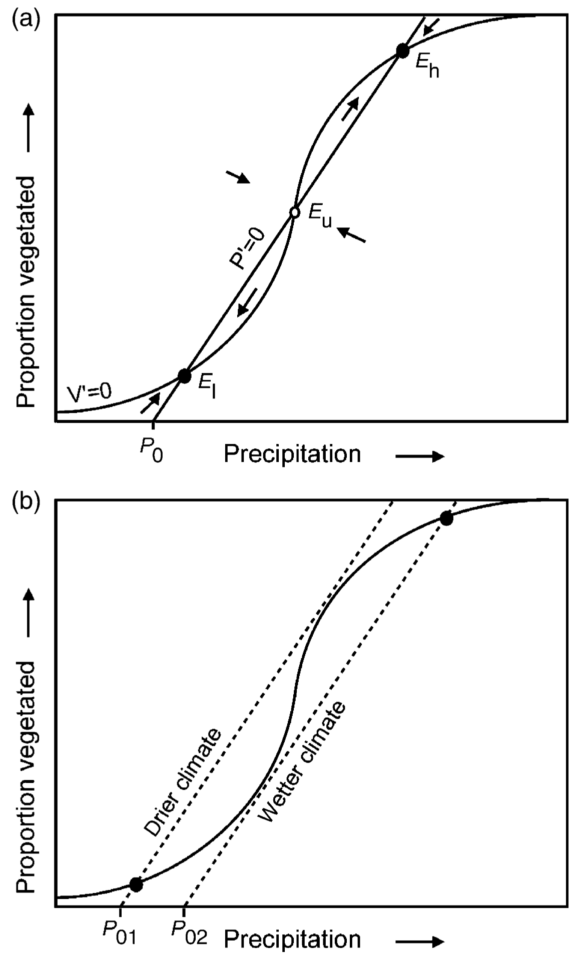

Figure 1.1: Schematic diagram showing the instability of the vegetation–rainfall system. "If vegetation enhances the amount of precipitation compared to the level in the absence of vegetation (P_0), the resulting tilted equilibrium precipitation ($P' = 0$) line may intersect with the vegetation equilibrium line ($V' = 0$) in three points (a). The arrows indicate the direction of change if the system is not in equilibrium and show that the intermediate equilibrium point (E_u) is an unstable saddle point, whereas the low and high biomass points (E_l and E_h) are stable. If general climatic conditions are sufficiently dry or wet, situations with single stable intersections may arise (b)." Note that the slope of the P' curve is not necessarily constant. Adapted from [40].

horizontally) over the land, providing moisture that rises in the warm environment over the land. This process produces more precipitation, reinforcing the vegetation state of the land.

Another mechanism by which vegetation feedback influences atmospheric circulation and rainfall is transpiration. Plants, particularly trees with deep root systems, are able to tap into groundwater during dry seasons. The moisture released into the atmosphere by transpiration enhances convection and thus rainfall. In fact, Myneni et al. (2007) have suggested that Amazonian trees transpire in anticipation of the onset of the rainy season, perhaps playing a role in the development of rainy season precipitation[3] [31]. The effects of transpiration also occur during wet periods, providing self-sustenance for a rainfall–vegetation feedback.

These mechanisms support the hypothesis of a positive feedback system in rainfall–vegetation interactions. Mathematically, such a positive feedback system has several important characteristics: the presence of one or more stable states; instability between stable states; abrupt transitions between stable states due to the instability; critical thresholds governing transitions between states; hysteresis, in that the critical threshold for transitioning to another state may not be the same as for returning to the original state; and decreased resilience (ability to recover from perturbations) near the transition thresholds. As an example, Levermann et al. [26] derive a model for monsoon

[3]This result is derived from the increased greenness measured by satellite instruments. There is debate as to whether this is caused by leaf reflectivity or transpiration.

dynamics similar to those described above but without the effect of vegetation[4]. This simplified model exhibits all of the above characteristics.

1.1.1 An example

It is useful to present here an example of feedback in a vegetation system to illustrate the general properties of such systems. Suppose we have a microclimatic two-state system of barren land and vegetated land. Suppose as well that there is a minimum amount of precipitation necessary for vegetation to colonize the barren land, that is, for vegetation to expand. We find that this colonization threshold is actually higher than the threshold of rainfall necessary to sustain the vegetated state. This is due to a number of factors, including shading of seeds by established plants [e.g., 37], collection of moisture by plants [13], and the mechanisms discussed above that enhance precipitation feedback. Therefore, even with precipitation below the colonization threshold, the vegetated state is self-sustaining. There is, however, a desertification threshold lower than the colonization threshold below which the vegetated state can no longer sustain itself and then collapses. This desertification threshold may be attained by severe and/or prolonged drought, depending on the specific characteristics of the vegetation and regional climate. This example is illustrated schematically in Figure 1.2.

[4]Indeed, the simple model derived in this study suggests that a similar model may exist for the rainfall–vegetation feedback mechanism; see section 4.2.

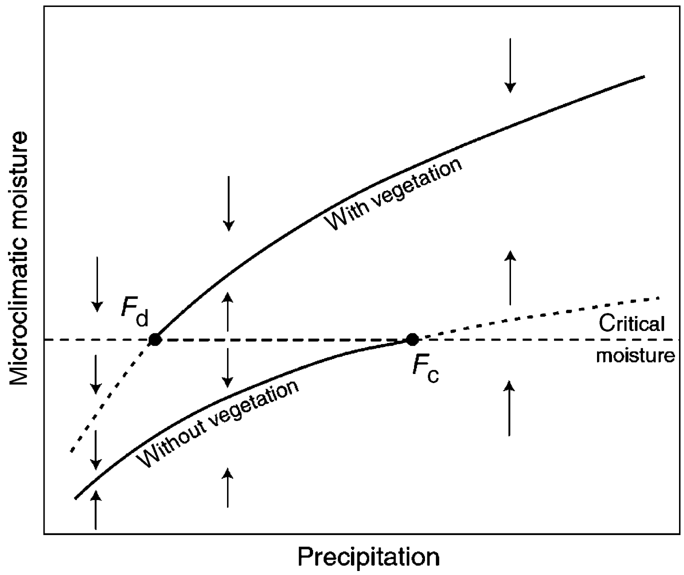

Figure 1.2: Schematic of the variable thresholds required for extinction and colonization of vegetation. “Microclimatic moisture conditions not only depend on precipitation, but are also enhanced by the presence of vegetation. This implies that in dry regions the critical moisture level may be too low to allow colonization by plants, whereas once vegetation is present, microsite moisture at the same precipitation level may be sufciently enhanced to allow growth and rejuvenation. As a result, the critical precipitation level for colonization of new areas (F_c) can be higher than the critical precipitation at which vegetation disappears (F_d).” Adapted from [40].

1.2 Empirical evidence for rainfall–vegetation feedback

Several recent studies have highlighted examples of vegetation changes, especially due to drought, and the characteristics of these changes. Two such examples are presented here: one of forest–woodland variation in New Mexico, and another of forest recovery in the Amazon after the 2005 drought. Studies demonstrating global patterns of vegetation states and possibly suggesting feedback systems are also presented.

Allen and Breshears (1998) [2] found that persistent drought in an upland region of New Mexico induced pronounced shifts in the ecotone (boundary between two types of ecosystems) between ponderosa pine trees and piñon–juniper shrubs. According to survey research, the study finds that the 1950s drought caused more than half of the pine forest in the studied tegion to be replaced by shrubs. Interestingly, while the mortality of the forest ceased once rainfall returned to normal values, the forest did not return to its previous land coverage. This finding supports the hysteresis model of abrupt vegetation changes, as the precipitation threshold for forest growth is larger than the precipitation necessary for forest sustenance. The authors conclude that the relatively short-duration 1950s drought had a permanent and lasting impact on the regional ecosystems by affecting such ecosystem properties as soil erosion.

Saatchi et al. (2013) [38] present a more recent study using satellite data to map changes in vegetation cover over the Amazon during and after a drought event in 2005. By analyzing rainfall and microwave scattering data (to determine the roughness of the land surface), the authors find that significant areas of the southwestern Amazon suffered losses of canopy cover. More remarkably, the results indicated that the canopy cover did not recover appreciably between 2005 and 2010 despite normal and even excessive rainfall in 2009. Like the previous study, this result suggests that a significant short climatological event can permanently alter the vegetation. Notably, the study also found that vegetation in regions with short dry seasons (and thus relatively low resistance to drought) was affected by rainfall deficits much sooner than vegetation in regions of long dry seasons.

While the above studies demonstrate the effects of abrupt changes in a vegetation feedback system on a small scale, there is also evidence for this vegetation model on global scales. An empirical survey by Hirota et al. (2011) [17] found evidence of three stable states of vegetation: forest (tree cover greater than 60%), woody savanna (tree cover 5–60%), and treeless (tree cover less than 5%). The frequency of occurrence of tree cover percentages plotted against average annual rainfall is trimodal, with peaks centered at 0%, 20%, and 80% tree cover, corresponding to treeless, savanna, and forest states, respectively. Because the regions in between those states are unstable, changes in precipitation do not cause gradual changes in tree cover, but rather a change in the probablility of being in each of the stable states. (A conceptual model of the instabilities involved in the tree cover–precipitation system is shown in Figure 1.3.) In addition, the study demonstrates that the resilience of an ecosystem decreases as it approaches a bifurcation point (unstable state). Certain areas of the Amazon rainforest are determined to be at high risk of shifting to savanna or treeless states.

Taken together, the above studies demonstrate that the dynamics of the vegetation feedback systems are complex and highly sensitive to short time0scale perturbations. One can conclude that these many complicated factors need further understanding in order to improve the quality of future

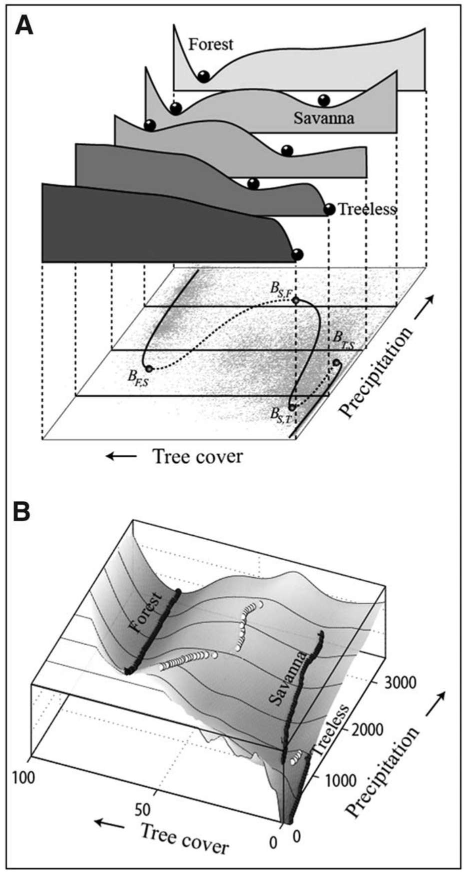

Figure 1.3: The treeless–savanna–forest vegetation stable state system showing the "relationship between the resilience of tropical forest, savanna, and treeless states and mean annual precipitation (mm yr^{-1}). (A) The tree cover data (percent, bottom plane) suggest a double instability. Stable states correspond to solid parts of the curve on the bottom plane and to minima in the stability landscapes. Unstable equilibria correspond to the dashed parts of the curve and to hilltops in the stability landscapes. At bifurcation points (B), stable equilibria disappear through collision with unstable equilibria. Resilience measured as the width of the basin of attraction around a stable state diminishes toward such bifurcation points. (B) Potential landscapes as computed directly from the data. Stable states (solid dots) are minima and the unstable equilibria (open dots) are maxima at a given precipitation level." Adapted from [17].

predictions of climate and land changes.

1.3 Details on the Congo region and its significance

The Congo Basin is of global climatological importance as it is the second largest rainforest in the world, after the Amazon. As such it is important in energy and carbon balances. Global patterns, especially the shift of the Intertropical Convergence Zone (ITCZ), cause a bimodal pattern of rainfall in the Congo region of equatorial Africa [32]. The ITCZ is the zone of convergence of trade winds near the equator that is primarily characterized by widespread convection and rainfall. The peak rainfall seasons in the Congo occur around October and March, when the ITCZ passes southward and northward over the region (this pattern is observed in this study, c.f. Figure 3.10). The volatility of the rainy seasons in the Congo is evident; the region has the greatest lightning frequency in the world.

While not well understood, a number of factors that affect precipitation in western Africa may also affect central equatorial Africa. These include sea surface temperatures in the Atlantic (which may follow large-scale oscillation patterns such as the El Niño–Southern Oscillation) and the strength of easterly jets. Generally, a weakening (strengthening) of mid-level African easterly jets and a strengthening (weakening) of the upper-level tropical easterly jet is associated with a decrease (increase) in rainfall in central equatorial Africa [32].

The vegetation of the Congo Basin is more drought-resistant than that of other rainforests because of the seasonality of rainfall. However, in the past decade, the Congo has experienced a sharp decrease in vegetation greenness. A recent study by Zhou et al. (2014) has shown that this decrease in greenness has been associated with decreases in rainfall, terrestrial water storage, biomass water content, and forest canopy [45]. Recent increases in surface temperatures and decreases in photosynthesis have also been observed. These concerning trends, and the possibility of vegetation shifts to deciduous forest or grassland due to lengthening dry seasons, are further reasons for a study of the Congo region.

1.4 Summary of motivations for this study

The aforementioned empirical examples of dynamical rainfall–vegetation feedback effects have highlighted the need to improve understanding of the characteristics of abrupt vegetation changes. While studies have suggested regional impacts of significant drought events on vegetation shifts, no work has yet developed a general pattern to suggest mechanisms for vegetation changes. We therefore attempt to characterize drought on a global scale to link drought with measurable changes in vegetation type. More specifically, we seek to answer two general questions:

1. On a global scale, what shifts in vegetation have been observed, and can these shifts be explained by drought? and

2. In a local study in the Congo Basin, does drought affect vegetation

changes, and can drought be identified by physical mechanisms in the climate system?

These questions are the topic of the following chapters.

Chapter 2

Data and Methods

2.1 Data

1. *Global Precipitation Climatology Project (GPCP)* [1, 44]

   The GPCP dataset is a comprehensive analysis of global precipitation from 1979–present. The data include station data, satellite measurements, and atmospheric soundings. Sources of error in the data include satellite measurement error, standardization of station data, and low spatial resolution. Maintained by the National Oceanic and Atmospheric Association (NOAA) Earth System Research Laboratory, the data are available in a 2.5-degree gridded format.

2. *Global Historical Climatology Network version 2 (GHCN)* [36]

   The GHCN v2 dataset is a comprehensive record of station rainfall with temporal range dependent on each station. The station latitude and longitude coordinates were added separately. Maintained by the NOAA National Climatic Data Center, the station data are thoroughly quality controlled and analyzed for extraneous errors.

3. *Tropical Rainfall Recording Mission (TRMM) 3B43 monthly*

   The TRMM satellite data span 1998–present and cover the globe from

50°S to 50°N. The particular dataset used is monthly averages of daily precipitation values. The TRMM satellite measures precipitation using a combination of microwave imagery, precipitation radar, visible and infrared imagery, radiation measurements, and lightning detection. Maintained by the National Aeronautics and Space Administration (NASA) and the Japan Aerospace Exploration Agency, the data are available in a 0.25-degree gridded format.

4. *University Corporation for Atmospheric Research Standardized Precipitation Index (UCAR–SPI)* [in 35]

   The UCAR–SPI dataset is a reconstruction of standardized precipitation index (SPI) that spans 1949–2012. 3-month, 6-month, and 12-month SPI are included. The data are derived from the University of East Anglia Climate Research Unit TS3.21. Maintained by NCAR/UCAR, the data are available in a 1-degree gridded format.

5. *Moderate-Resolution Imaging Spectrometer (MODIS) MCD12A1 Combined Land Type* [9]

   The Moderate-Resolution Imaging Spectrometer satellites (Terra and Aqua) measure land types via greenness and near-infrared reflectivity. The Combined Land Type includes the percentage of each grid point measured to be any of a list of land types (see Table 2.2). The MCD12A1 data are a yearly composite of average land type and are available from 2001–2012. Maintained by NASA and the Goddard Space Flight Center, the data are available in a 0.05-degree gridded format.

6. *MODIS MCD15A3 Combined Leaf Area Index* [9]

   From the same satellite data as above, the Leaf Area Index (LAI) measures the vertical areal coverage of leaves. The range of LAI is from 0 to 7, with 7 corresponding to crossing an average of 7 leaves in a vertical profile. The data are a 4-day composite and are available from July 4 2002–present. The data are calculated by determining the fraction of photosynthetically active radiation absorbed by vegetation but are subject to interference such as clouds and aerosols. There are also over-estimate anomalies for very green vegetation such as new leaves. Maintained by NASA and the Goddard Space Flight Center, the data are available in a 1-km^2 resolution tiled format with sinusoidal projection.

7. *European Centre for Mid-Range Weather Forecasting (ECMWF) Re-Analysis (ERA-Interim)* [11]

   The ECMWF reanalysis data are constructed from models supplemented with observational data for the years 1979–present. Caveats involved with reanalysis data are primarily model uncertainties. Maintained by the ECMWF, the ERA-Interim data are available in a 1.5-degree gridded format.

2.2 Methodology

In this section we describe some of the methods used in obtaining results, including the physical parameters used, the calculations performed, and

Moisture category	SPI index
exceptionally moist	≥ 2
extremely moist	1.60 to 1.99
very moist	1.30 to 1.59
moderately moist	0.80 to 1.29
abnormally moist	0.51 to 0.79
near normal	–0.50 to 0.50
abnormally dry	–0.79 to –0.51
moderately dry	–1.29 to –0.80
severely dry	–1.59 to –1.30
extremely dry	–1.99 to –1.60
exceptionally dry	≤ -2

Table 2.1: Moisture levels for the standardized precipitation index, as used by the National Weather Service. Drought thresholds are negative.

the coding used in the process.

2.2.1 The standardized precipitation index (SPI)

The indicator used for determining drought conditions is the standardized precipitation index (SPI). Originally developed by McKee et al. (1993) [29], the SPI is a statistical measure of the deviation of precipitation from the climatological mean on a month-to-month basis. Typically, the SPI is measured on a 3-month, 6-month, or 12-month time scale, indicating the degree to which the past 3 (or 6, or 12) months have been dry or wet compared to the norm. Because the SPI takes only precipitation into account in determining drought (unlike some other widely used indicators such as the Palmer Drought Severity Index which include factors such as soil moisture), it has some advantages and disadvantages. On the plus side, the indicator is not affected by soil moisture or vegetation feedback which may unusually amplify or dampen the variability. However, because other factors are not included, the SPI may not accurately reflect the impacts of deviations such as abnormally long dry seasons which may have much more of an effect on the vegetation than on rainfall deviations. Table 2.1 tabulates the categories of drought based on SPI. We will refer primarily to "severe" and "extreme" drought.

2.2.2 General definitions of drought

In order to standardize the measurements of drought, we need to define a drought as used in this study. Drought is measured on a monthly time scale, with SPI-6 (6-month SPI) and SPI-12 (12-month SPI) the primary indicators. The characteristics of drought used were average drought length and total months of drought. The average drought length is determined by measuring a drought event as follows: a drought event is initiated when the SPI exceeds a given threshold, and continues *until the SPI returns to positive values – normal to wet conditions.* The average length of all the drought events within the time series is then calculated. The total months of drought is a count of all months within a given time series that are classified as drought according to the scheme just described.

IGBP land classification	Simplified vegetation scheme
Water	—
Evergreen needleleaf forest	Forest
Evergreen broadleaf forest	Forest
Deciduous needleleaf forest	Forest
Deciduous broadleaf forest	Forest
Mixed forest	Forest
Closed shrublands	Desert
Open shrublands	Desert
Woody savannas	Grassland
Savannas	Grassland
Grasslands	Grassland
Permanent wetlands	—
Croplands	—
Urban and built-up	—
Cropland/Natural vegetation mosaic	—
Snow and ice	—
Barren or sparsely vegetated	Desert

Table 2.2: Land classifications of the IGBP scheme as used in the MODIS land type data [9], and the corresponding vegetation type used in the three-type simplified vegetation scheme in this study. Land types marked "—" are not included in the simplified scheme.

2.2.3 Assessing land changes

The MODIS data used for land type classify each grid point with a percentage cover of each of 16 classifications according to International Geosphere-Biosphere Programme (IGBP) Plant Functional Classifications. For example, a 0.05-degree grid point may be 50% evergreen broadleaf forest, 30% mixed forest, and 20% woody savanna. In this study, the 16 classifications are simplified to the three broader types – desert, grassland, and forest – eliminating points with water, wetlands, urban development, and croplands (see Table 2.2). We use this simplified vegetation scheme to more accurately characterize changes between the three broad stable states as determined by Hirota et al. [17]. In the case study for the Congo region, we also plot time series of the vegetation type and of the leaf area index (LAI) in order to gain a more precise understanding of the changes occurring in the specific region.

2.2.4 Coding and resources employed in calculations

Nearly all of the calculations presented were coded using the National Center for Atmospheric Research (NCAR) Command Language (NCAR Command Language or NCL), running on a CentOS Unix system. The SPI values for the GPCP precipitation dataset were calculated in NCL (these data are in addition to the UCAR–SPI data) as were all of the drought metrics described in the previous section. The atmospheric circulation calculations for the analysis of drought in the Congo basin were performed using the Grids Analysis and Display System (GrADS).

Graphics were produced primarily in NCL. Some of the graphics (for the atmospheric circulation of the Congo) were produced in GrADS.

Chapter 3

Results and Discussions

In this chapter we present the results of our calculations and analysis. The chapter is organized into three sections: observations on a global scale, observations in the tropics (15°N – 35°S), and local study of the Congo basin. At the end of each section we present the implications of current findings.

3.1 Global observations

3.1.1 Global precipitation

Climatological rainfall is plotted in Figure 3.1. This figure is mainly included as a reference for future results that indicate land types and land type changes. Precipitation data were also obtained from the TRMM satellite for the tropical regions, and the two sources of data are in good agreement for climatological rainfall (see Figure A1.1).

3.1.2 Land changes over the past decade

We are interested in measuring how the three broad vegetation states – forest, grassland, and desert – have varied over the past 10 or so years of available data, with the goal of finding evidence for vegetation stable states and abrupt changes. (Figure A1.2 shows global predominant land types for

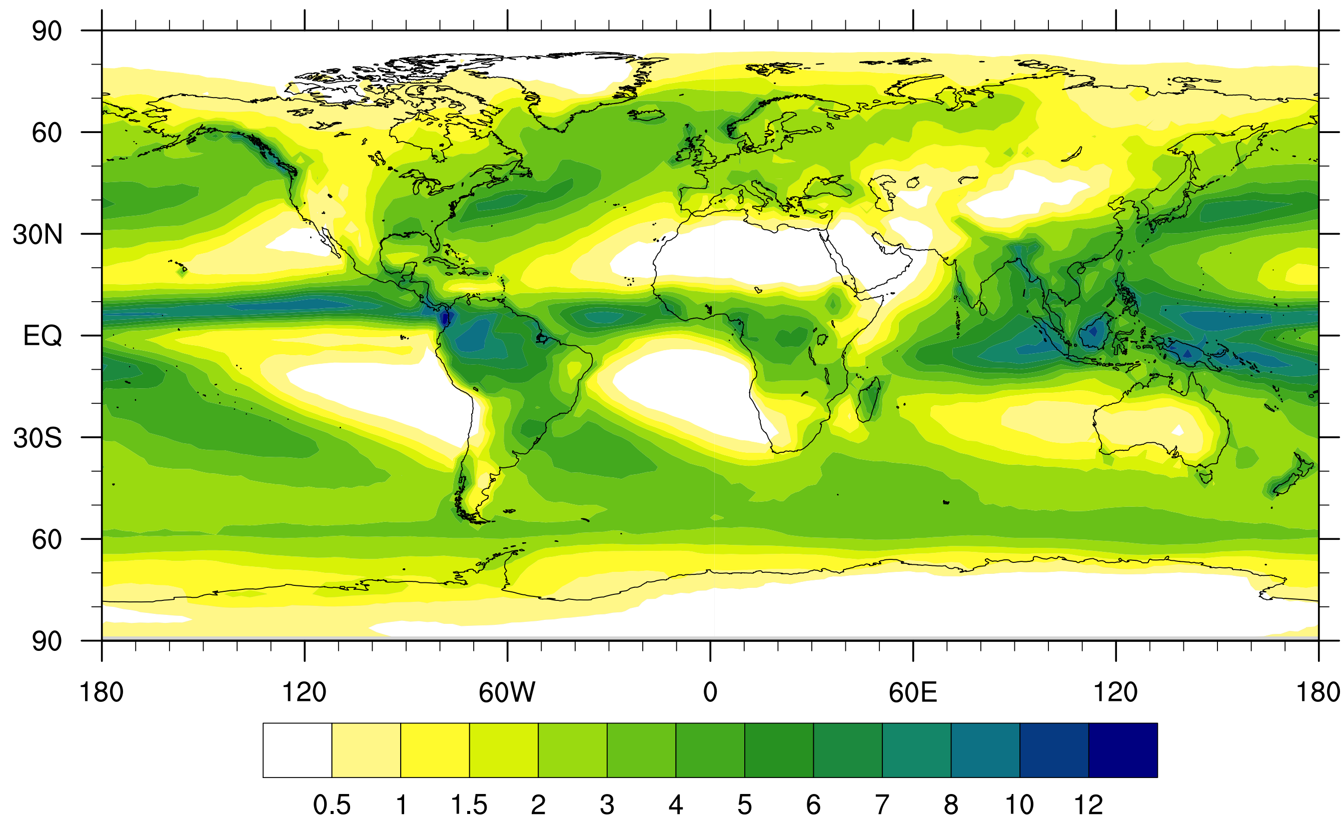

Figure 3.1: Average climatological rainfall from 1979–2012 as measured in the GPCP dataset. Units are mm day^{-1}.

reference.)

In order to assess land changes, we use trends in land percentage over the past 6 years (2007–2012) and the past 12 years (2001–2012). The trends, which are employed because the land use data spans from 2001–2012 (and so the time range is more restricted than for rainfall data), highlight where land changes have occurred and to what extent. The changes in land type are shown in Figures 3.2 and 3.3 for 12 years and 6 years, respectively. Of note are several regions of high land variability, including the Sahel region of Africa, much of Australia, southern Africa, and parts of central South America. In these regions, there is a dominant interchange between two specific vegetation types. For example, in the Sahel, there is an increase in grassland at the expense of desert, while in Australia, desert increases while grassland decreases.

3.1.3 Global measurements of drought

To investigate whether there is a link between drought and vegetation variability we analyzed drought on a global scale using the UCAR–SPI data. Using the drought definition of the previous chapter, total months of drought and the average length of drought events were calculated for each 1-degree grid point. These results are shown in Figures 3.4 and 3.5 respectively for a drought threshold of "extreme" (SPI < -1.59). The former result is obtained using 6-month SPI in order to more accurately count all drought events, whereas the latter was obtained using the 12-month index to capture events of prolonged drought. The threshold of "extreme" was used to isolate significant drought

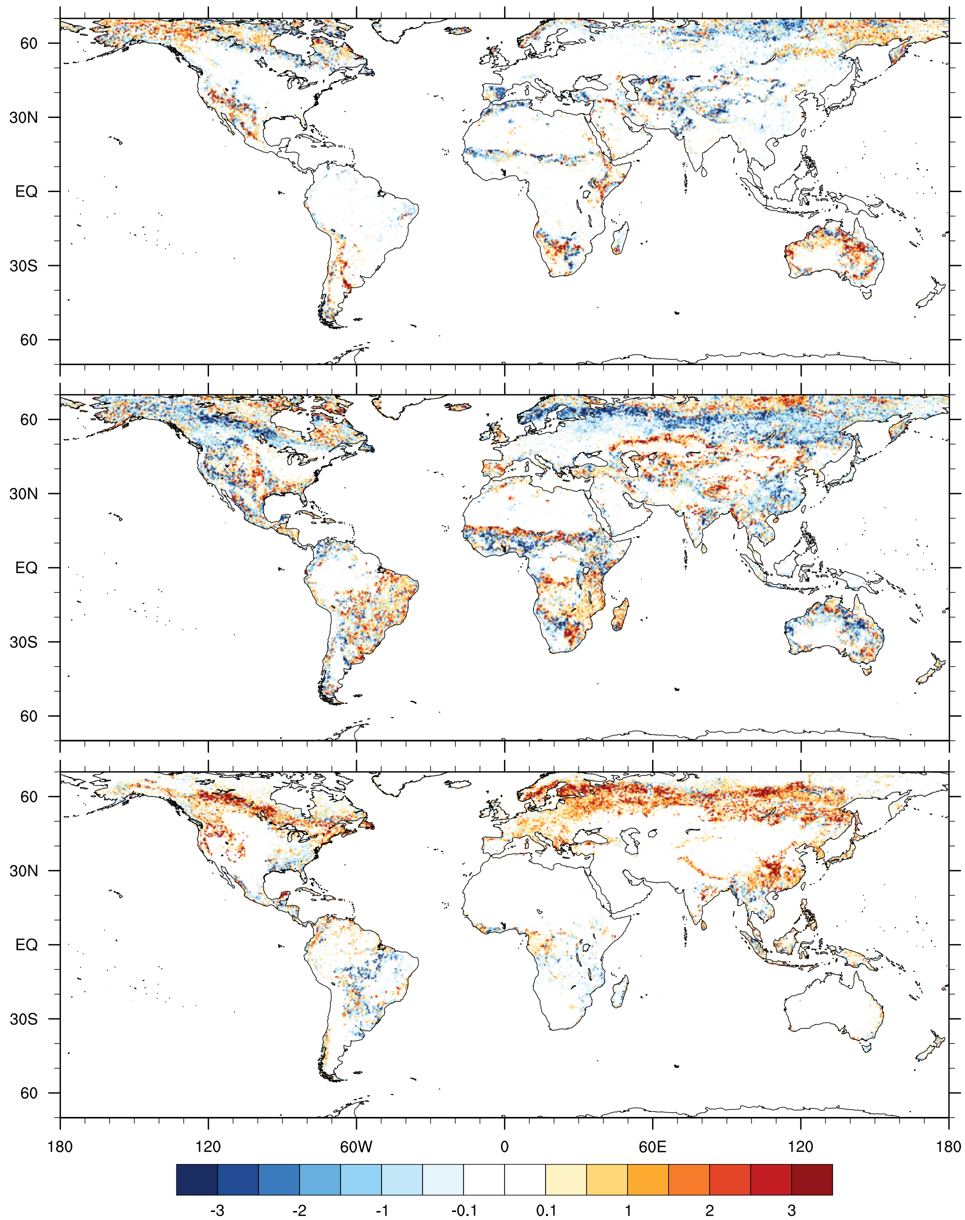

Figure 3.2: Average yearly rate of change in percentage cover of desert (top), grassland (middle), and forest (bottom) from 2001–2012, as measured from MODIS land type data. Note regions of variation and interchange between land types, such as the Sahel and Australia.

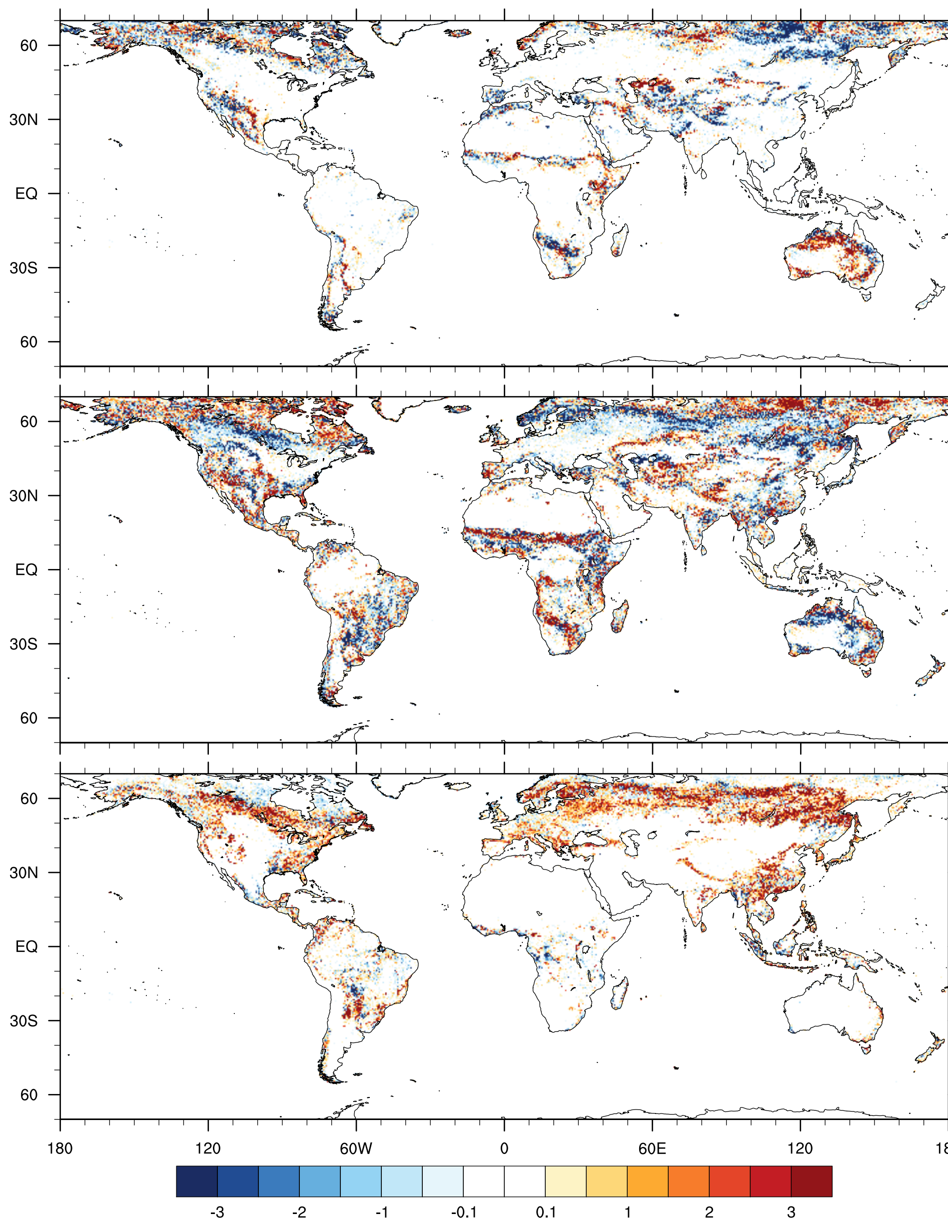

Figure 3.3: Average yearly rate of change in percentage cover of desert (top), grassland (middle), and forest (bottom) from 2007–2012, as measured from MODIS land type data. Note regions of variation and interchange between land types, such as the Sahel, southern Africa, and Australia.

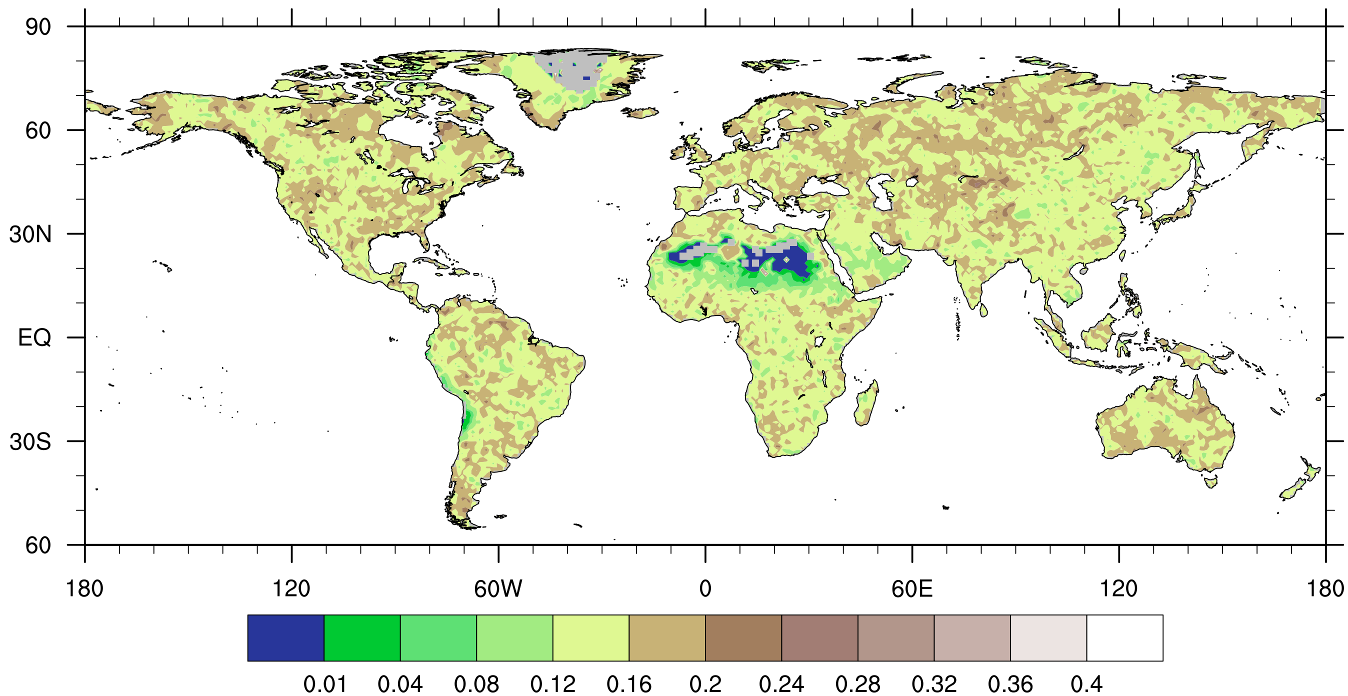

Figure 3.4: Fractional number of months from 1949–2012 in extreme (SPI < -1.59) drought conditions, as derived from 12-month SPI in the UCAR-SPI data. Apart from barren regions such as the Sahara that do not experience drought, there is little variation in occurrence of drought.

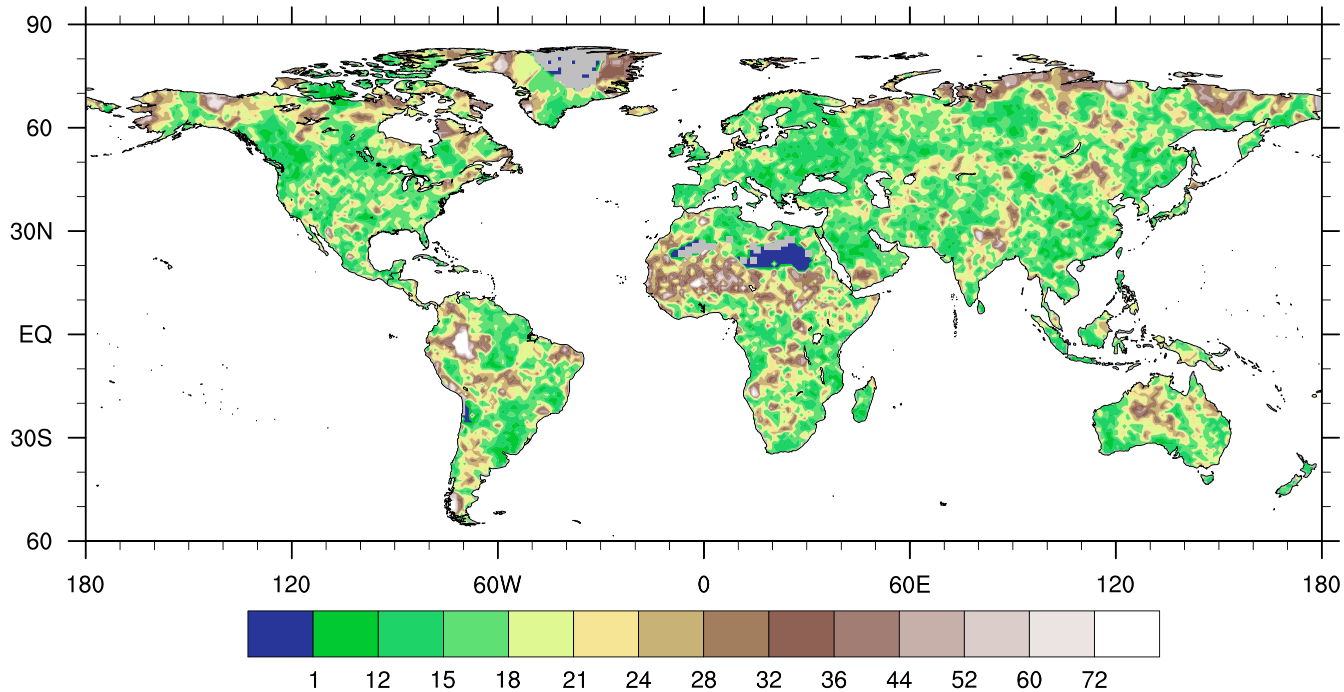

Figure 3.5: Average length of extreme (SPI < −1.59) drought events (in months) from 1949–2012, as derived from 12-month SPI in the UCAR-SPI data. Note several regions of high occurrence of long drought, especially the Sahel and central Australia.

events.

3.1.4 Discussion of global results

Examination of the land type change and drought maps reveals some features of interest. First, as mentioned above, areas with significant changes in land type show opposite changes in another land type, and more specifically in a neighboring land type on the rainfall scale. For example, the Sahel shows an increase in grassland while desert decreases and the opposite is true of much of Australia. Similarly, there is a clear interchange between grassland and forest in central South America that is spatially exact. These observations suggest that the observed vegetation changes are induced by rainfall changes rather than external factors such as deforestation, providing evidence for stable states in the rainfall–vegetation feedback system. Unstable states are the regions that are prone to vegetation changes (the Sahel, in particular, stands out). It should briefly be noted that areas around the Arctic circle show an increase in forest cover that is not entirely balanced by loss of grassland, or even desert. This is likely due to a loss in barren or ice-covered tundra (possibly due to warming temperatures) that is not represented in the simplified vegetation scheme[1].

The comparison between the maps of drought and those of land changes is also interesting. Regions with highly variable vegetation appear to also be prone to long drought events (Figure 3.5). These regions are not, however,

[1]Later results will focus on land changes in the tropics, where temperature changes are not as important.

prone to more drought (Figure 3.4). This suggests that drought alone (even extreme drought) is not an adequate explanation for vegetation changes; it is long drought events that have most effect on vegetation. There is, however, no significant statistical correlation between land variability (measured as the sum of the standard deviations in forest, grassland, and desert) and average drought length.

3.2 Observations for the tropics

In the tropics (15°N – 35°S), the effects of temperature and synoptic weather systems on vegetation are greatly reduced. Thus, vegetation types and changes are primarily determined by precipitation and variations in annual precipitation cycles. By analyzing the tropics, then, we eliminate extraneous factors that could impact land changes.

3.2.1 Rainfall changes in the last decade

Figure 3.6 shows the change in precipitation over the last decade from 2002–2012 derived from high-resolution TRMM rainfall data. Some of the features in land changes seem to be represented as rainfall variations; for example, there is an increase in rainfall corresponding to the increase in grassland in the Sahel and a decrease in rainfall over the Congo area corresponding to a loss of forest. In other places, however, no correlation is observed. In Australia, a decrease in rainfall might explain the increase in desert in the western and northern parts of the country, but the eastern and central parts of the country

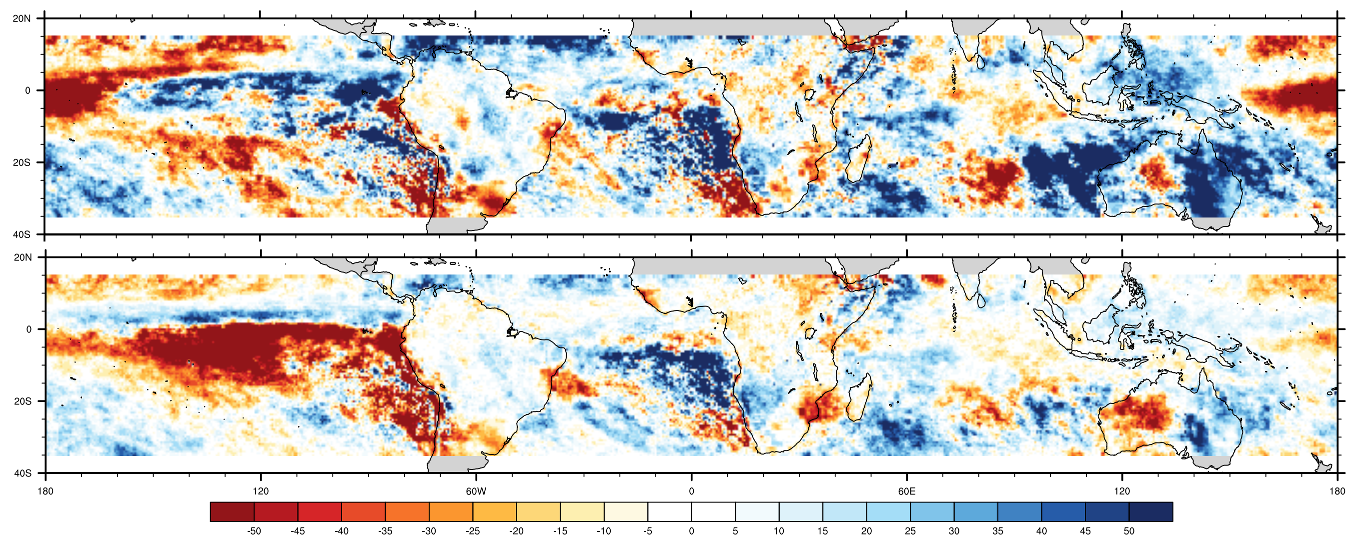

Figure 3.6: Percent changes in precipitation, 2012 minus 2002. The precipitation amounts for 2012 and 2002 were determined by averaging the previous two years (top) and five years (bottom). The changes in rainfall do not necessarily correlate with changes in vegetation (Figures 3.2 and 3.3.

actually experienced increases in rainfall.

3.2.2 Occurence of long drought related to annual precipitation

The three stable states (desert, savanna, and forest) identified by Hirota et al. [17] exist as a function of average annual precipitation. We expect to find that annual precipitation of less than 500 mm (approximately) is stably desert, precipitation of 1000–2500 mm is stably grassland, and precipitation of more that 2500 mm is stably forest (see [41] for quantifications). Thus, one might expect to find unstable states in transition values of precipitation, around 500–1000 mm and 2500 mm. If drought is a mechanism by which abrupt changes occur, these unstable regions are likely to see a high occurrence of long drought (as was observed in the global results from before).

Figures 3.7 and 3.8 show the occurrence of long drought as a function of rainfall for GHCN station data and UCAR–SPI grid points, respectively. For the station data, the plot of all stations does not show significant characteristics. However, separating the stations by the dominant land type reveals an increase in the occurrence of long drought at the extremes of the rainfall spectrum for all three land types, with the exception of the low end of desert (where drought does not usually occur for lack of any climatological precipitation). The same result is obtained using the UCAR–SPI drought data with TRMM annual rainfall measurements. The results are thus robust across two different sets of data, increasing confidence.

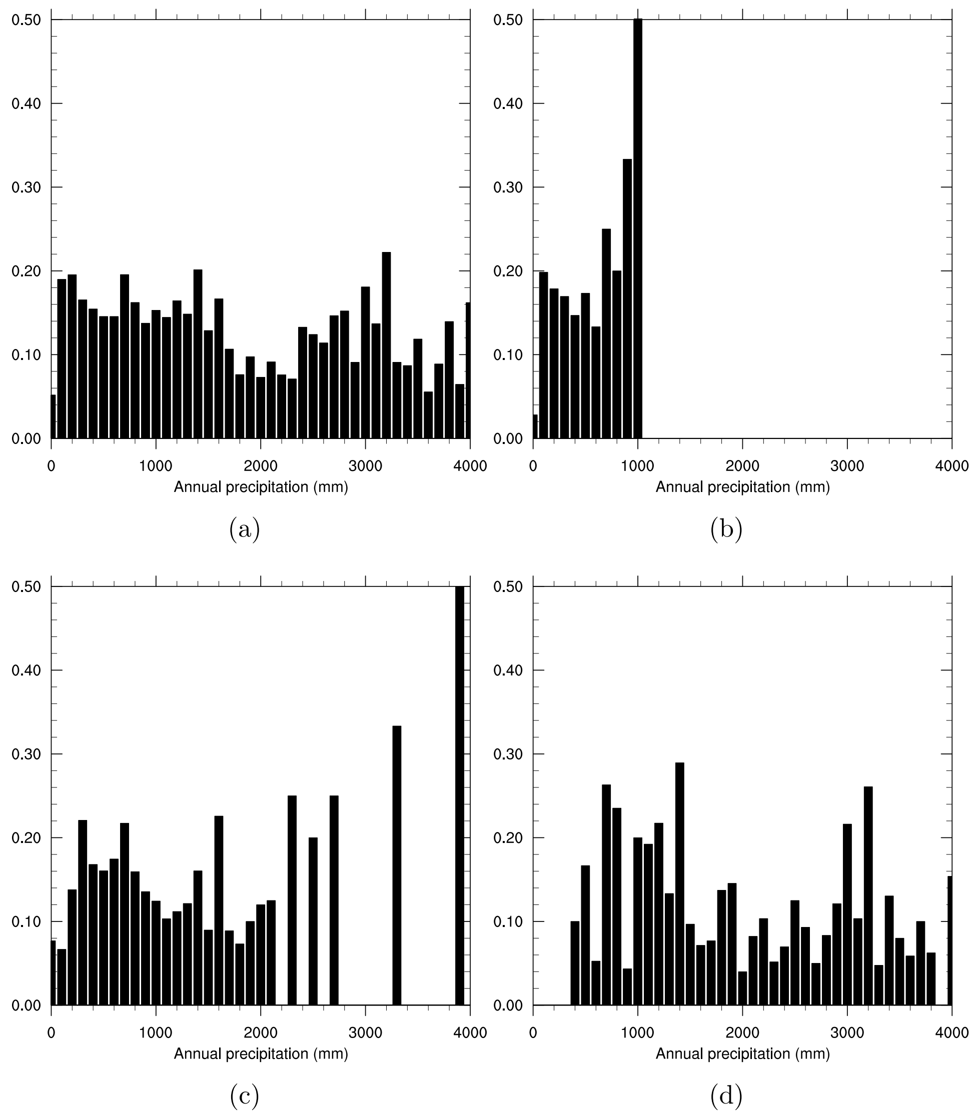

Figure 3.7: Spatial occurrence of long drought in the tropics (15°N – 35°S) using GHCN station data for (a) all stations, (b) stations in desert regions, (c) grassland regions, and (d) forest regions. The bars represent the percentage of stations within each 100 mm yr^{-1} rainfall tier that have an average SPI12-derived drought length of at least 18 months. Note that long drought occurs more towards the ends of the rainfall spectrum within each land type, except for the low end of desert. The dominant land types were determined from MODIS data for the year 2012.

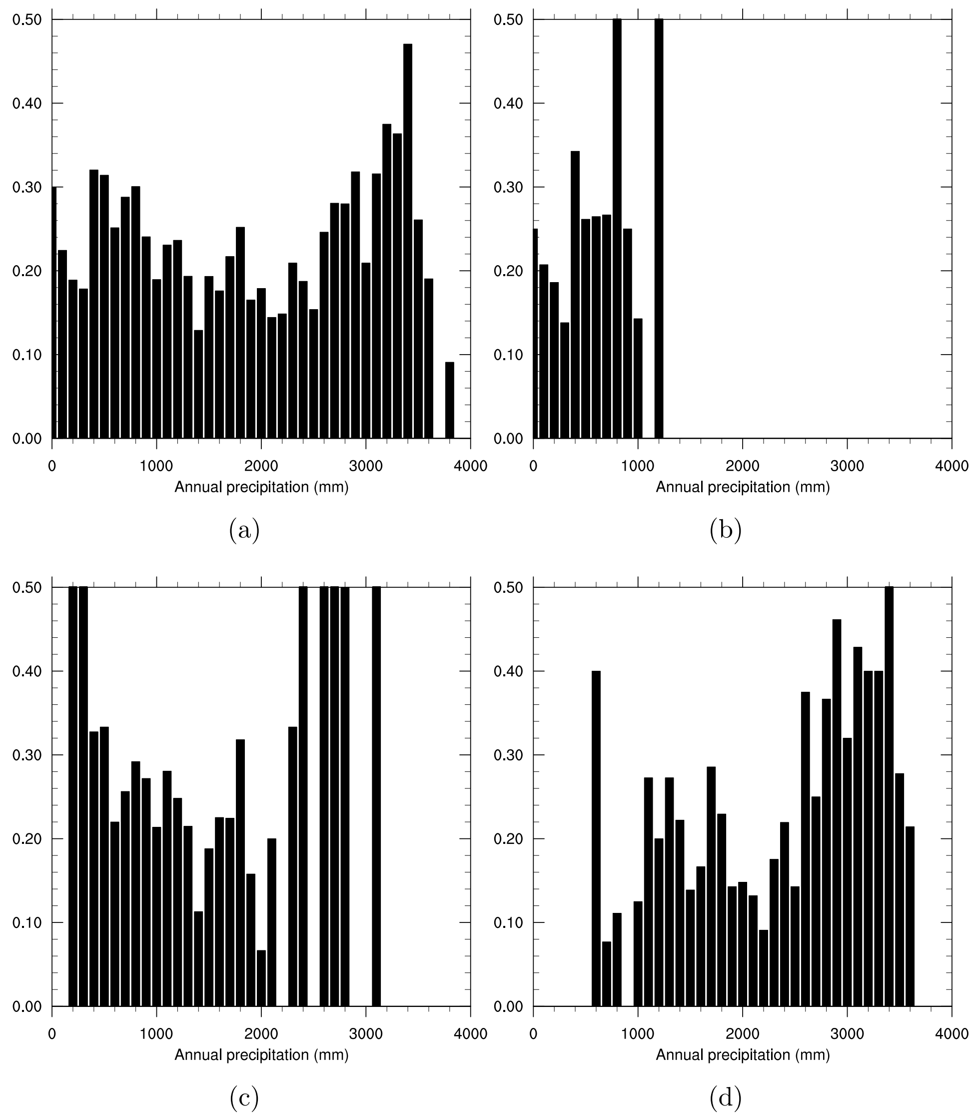

Figure 3.8: Spatial occurrence of long drought in the tropics (15°N – 35°S) using 1-degree grids from UCAR-SPI data for (a) all grid points, (b) grid points with majority desert cover, (c) grassland points, and (d) forest points. The bars represent the percentage of grid points within each 100 mm yr^{-1} rainfall tier that have an average SPI12-derived drought length of at least 24 months. Note that long drought occurs more towards the ends of the rainfall spectrum within each land type, except for the low end of desert. The dominant land types were determined from MODIS data for the year 2012 and the precipitation was determined from TRMM data.

3.2.3 Discussion of tropics results

We have shown that within each land type, the occurrence of long drought increases towards the extremes of the rainfall range in which the land type exists. This suggests that these regions of instability are affected by long drought, and by extension that long drought may be a mechanism of abrupt changes between vegetation types. As an example, suppose there exists a region of grassland with an average annual rainfall at the low end of the spectrum. This region is likely prone to a dieback resulting in a change in vegetation to desert. However, a desert region with high annual rainfall will likely green into grassland or savanna, and therefore it is unstable as well. In both instances, there is an increased susceptibility to long drought demonstrated in the above results, and so drought is likely a mechanism for the change between grassland and desert that occurs in these regions. Incidentally, the peak of long drought occurrence for dry grassland is at a lower value of precipitation than that for wet desert. Hence there is also evidence of positive feedback hysteresis in that precipitation generally must be higher for colonization of vegetation (wet desert to become grassland) than for dieback (dry grassland to become desert). The results therefore provide evidence for the trimodal stable state model and for hysteresis at points of instability, as discussed in the introductory chapter.

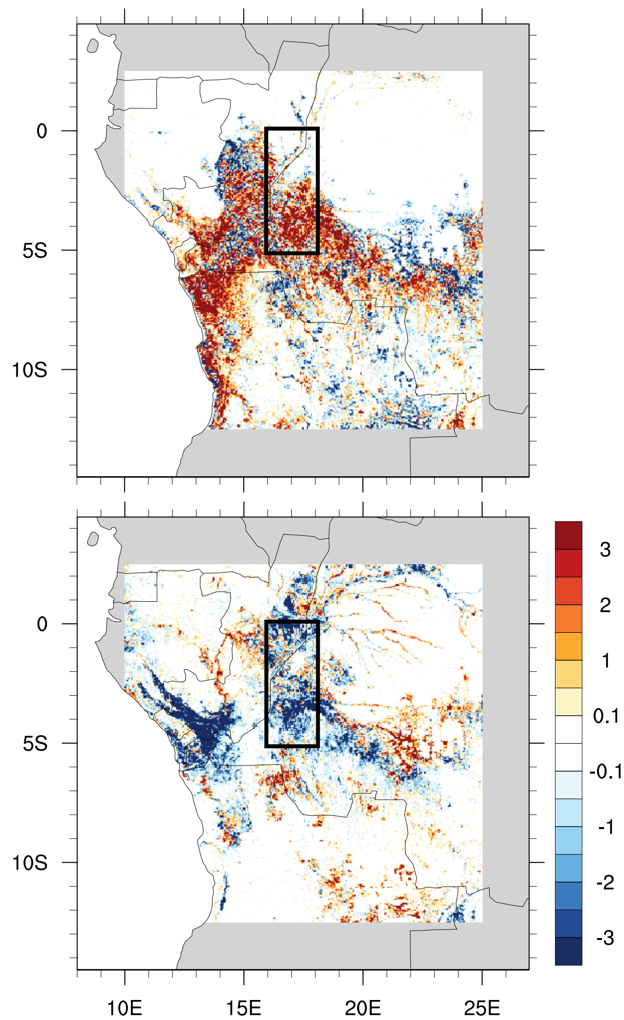

Figure 3.9: Average yearly rate of change in percentage cover of grassland (top) and forest (bottom) from 2007–2012, as measured from MODIS land type data. The black rectangles indicate the region of interest for this study. Note the decrease in forest cover within the indicated region.

3.3 Observations for the Congo Basin

As indicated in the first chapter, we analyze rainfall, drought, and vegetation characteristics for a localized region in the Congo Basin to determine whether there are measurable local indicators for the impacts of drought. The selected region covers 0°–5°S, 16°–18°E. This region is chosen on the basis of the land type changes map, as it has experienced a significant loss of forest in the past six years (Figure 3.9). While not an equivalent result, this decline in forest may be linked to observed declines in greenness and productivity of the Congolese rainforest [45].

3.3.1 Time series of drought and vegetation parameters

Time series for several parameters for drought in the Congo region are graphed in Figure 3.10. These time series were obtained by spatially averaging 6-month SPI from the UCAR–SPI data, rainfall and rainfall anomaly from TRMM data, LAI from MODIS LAI data, and land type percentage from MODIS land type data.

Several observations can be made in the context of these time series. First, the percentage of forest cover and percentage of grassland cover are clearly negatively correlated; the correlation coefficient is –0.43. This means that changes observed in the vegetation are an interchange between the two aforementioned vegetation types and not changes due to other influences (such as anthropogenic deforestation for agricultural purposes).

Second, there is agreement between the different SPI and rainfall anomaly

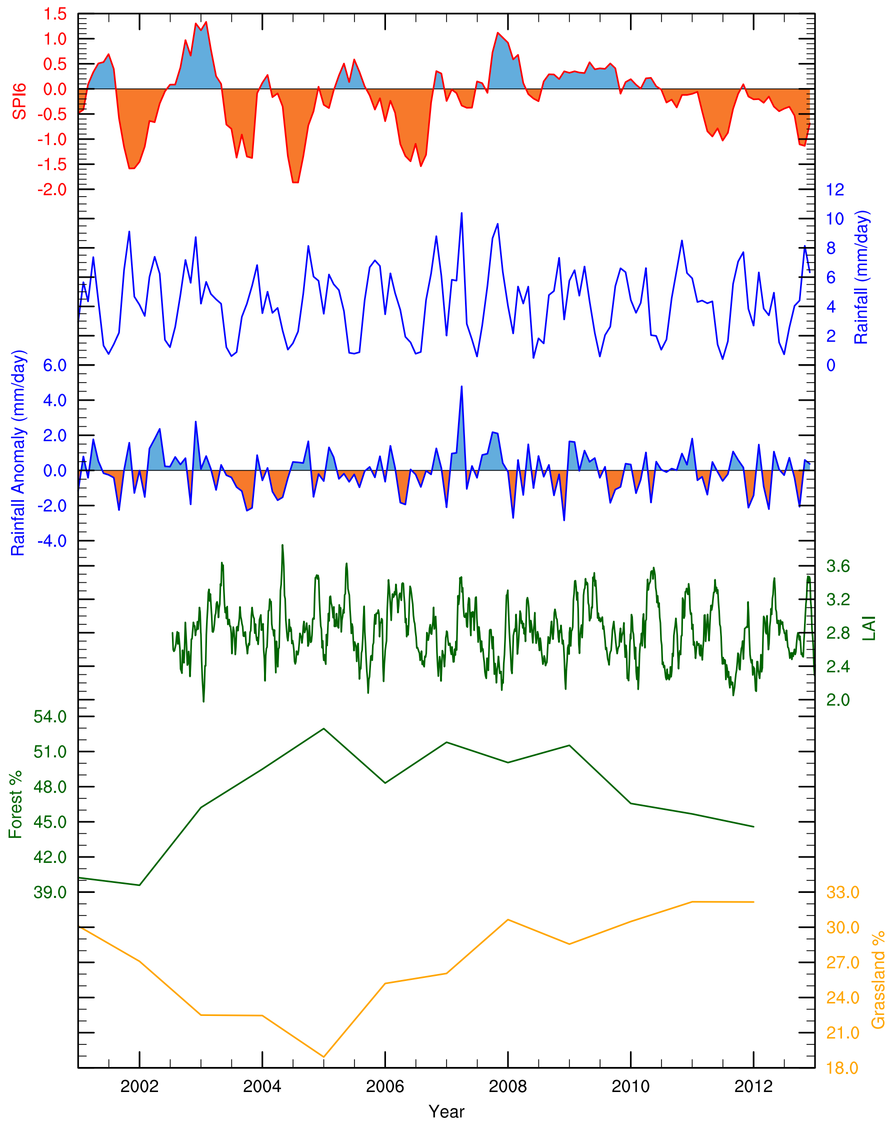

Figure 3.10: Time series for physical parameters of drought and vegetation change in the Congo Basin region (0°–5°S, 16°–18°E). From top to bottom: 6-month SPI from UCAR–SPI data, rainfall from TRMM data, rainfall anomaly from TRMM data, LAI from MODIS LAI data, percentage forest cover from MODIS land type data, percentage grassland cover from MODIS land type data. The negative correlation (–0.43) between the forest and grassland indicates an interchange between the vegetation types. The SPI and rainfall anomaly are somewhat in agreement (see Figure 3.11).

datasets. The correlation between these two time series is shown in Figure 3.11. Because the 6-month SPI measures drought over the previous 6 months, one would expect the best correlation to be with a lag of 6 months (i.e., the rainfall deficit or surplus leads the drought index by 6 months); however, the highest correlation coefficient occurs for a 2-3 month lag.

Third, there is no clear correlation between drought events and forest loss, and likewise between wet periods and forest increase. From 2001–2007, there are four significant drought events with SPI6 below -1.00, but forest cover actually shows a general increase[2]. The longer drought event from mid 2010 through 2012 does, however, correspond with the loss of forest as observed in the time series and in Figure 3.9. To better understand this dynamic between drought and vegetation, further work needs to be done on the correlation between drought events and forest (or grassland) loss on a regional scale such as this.

And finally, the leaf area index does not show any significant correspondence to the percentage of forest cover, but rather seems to have a more seasonal cycle. The LAI data should be further analyzed by removing seaonality, for example.

[2]It should be noted that, because the time scale for land percentage is in years, each point in the land plots corresponds to the average over the entire year following it. Thus it is not always clear where a correlation may exist.

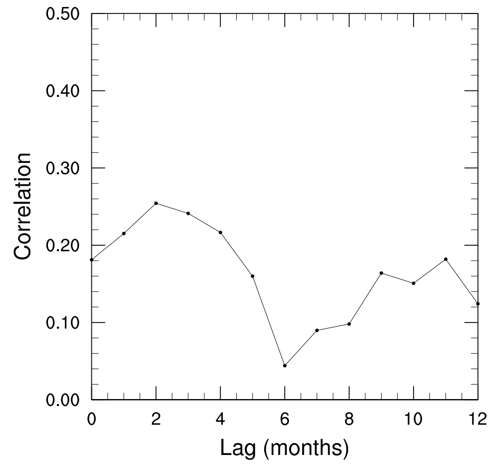

Figure 3.11: Correlation between 6-month SPI and rainfall anomaly as a function of lag (in months) for the select region of the Congo Basin, intended to verify the agreement between the UCAR–SPI and TRMM data. The lag indicates by how many months the precipitation anomaly leads the SPI change.

3.3.2 Atmospheric circulation in drought periods

As an extension to this regional study, we are also interested in analyzing the patterns of atmospheric circulation during periods of drought. Findings from this analysis may potentially be used to identify imminent drought conditions in future events, particularly for the little-studied Congo region.

As of the time of this writing, only preliminary results have been obtained. Drought years were identified by analyzing the UCAR–SPI data for the select region of the Congo and finding years where the austral (southern hemisphere) fall months of March–April–May (MAM) were abnormally dry. The threshold used for drought was an average of –1.00 or less across the region. The drought years during the period spanning the range of ERA-Interim data (1979–2012) were determined to be 1984, 1992, 1997, and 2004. The at-

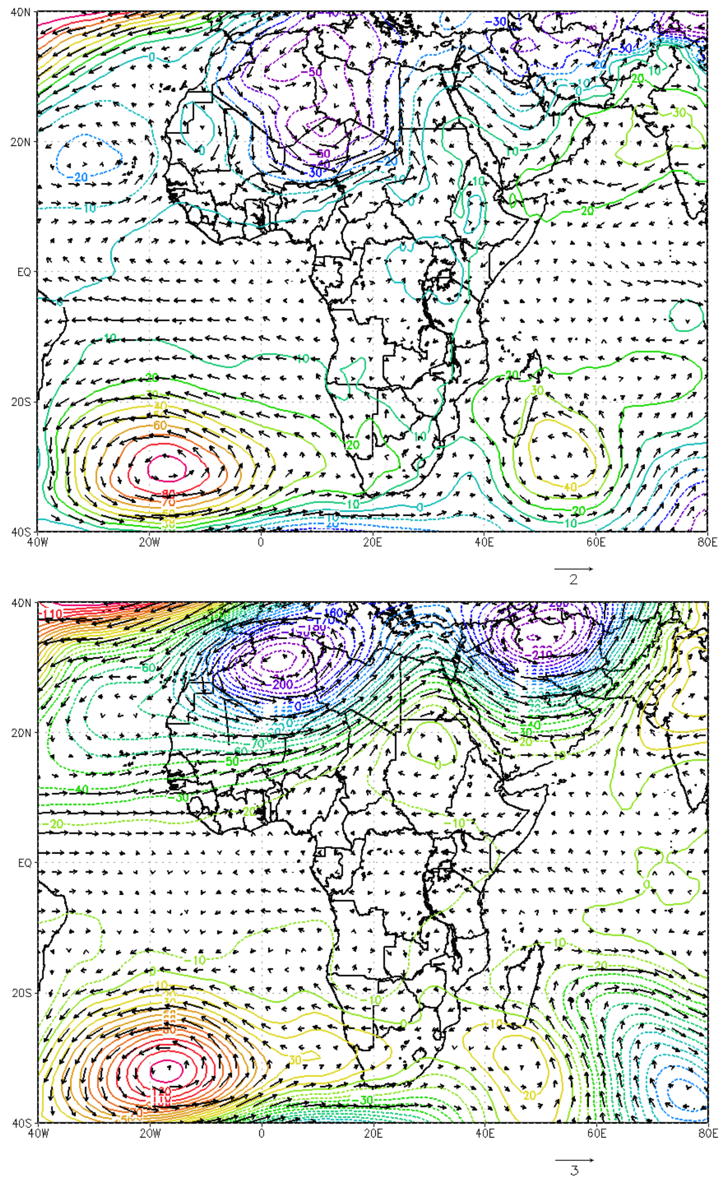

Figure 3.12: Geopotential height (contours) and horizontal wind (vectors) anomalies in March–April–May (MAM) for a composite of four drought years (1984, 1992, 1997, and 2004) minus the climatological mean, derived from ERA-Interim data. The top figure is at a pressure height of 850 hPa and the bottom is for 500 hPa. The drought years were selected for abnormally dry (SPI < -1.00) MAM rainy seasons. Note especially the region of high pressure in the southern Atlantic ocean and the westerly wind anomalies just north of the equator. Units are geopotential height meters (gpm) for height contours and m s^{-1} for wind vectors.

mospheric parameters for these four drought years were then averaged and compared to the climatological average for the entire time period. Figure 3.12 shows the wind and geopotential height[3] anomalies for the four drought years.

There are a few features of note in this first analysis. One is the strong high pressure anomaly in the southern Atlantic. Previous studies have linked a strong south Atlantic high with a decrease in rainfall over central tropical Africa [43]. The other main feature is the westerly flow anomaly north of the equator seen especially at the 500 hPa level (see Figure 3.12). Climatologically, the equatorial region from about 10°N–10°S is dominated by easterly (from the east) flow, while the mid-latitudes north and south of about 15° are dominated by westerly flow. Thus the results suggest that in drought years there is an increase in westerly flow across the Sahel and a slight decrease in easterly flow just north of the equator at the 500 hPa level. This combination may cause a northward shift of convective precipitation to north of the Congo Basin [e.g., 43]. Indeed, as mentioned in the Introduction, the weakened mid-level easterly flow correlates to an increase in rainfall in the Sahel region and a decrease in

[3] A note for readers not familiar with geopotential height in the atmosphere: For numerous practical reasons, including the cancellation of density factors in equations governing atmospheric fluid dynamics, the atmosphere is analyzed on pressure surfaces, where the pressure rather than the absolute height from the surface is held constant. Since pressure is a measure of the weight of air above, pressure always decreases with height, so pressure levels "descend" (approximately exponentially) with increasing height z above the surface. Areas of higher than normal pressure will have a pressure surface at a higher z, and areas of lower pressure will have a lower z. Thus a higher pressure at a fixed height is qualitatively equivalent to a higher height at a fixed pressure. The geopotential height is a more precise measure of z that integrates over changes in the gravitational constant g as z changes. In practice, variations in g are small due to the thinness of the atmosphere and so g is usually set as the constant $g_0 = 9.81$ m s^{-2} at the surface.

rainfall in equatorial Africa [32]. It should be noted that no major features are observed close to the region of interest. As a result, any mechanism for a decrease in rainfall would likely occur in larger-scale circulation.

3.3.3 Discussion of Congo Basin results

In this local study in the Congo Basin we have shown several features of the rainfall–drought–vegetation system in detail. Notably, the forest cover and grassland cover are negatively correlated, indicating that vegetation shifts between these two land types do occur. Additionally, the years 2007–2012 show both an extended drought and forest loss, suggesting that further investigation is needed to determine whether drought induces such a vegetation change. No general correlation was identified between drought and vegetation change in the available data.

In terms of atmospheric circulation patterns affecting drought, we observe a strong south Atlantic anticyclone and a westerly flow anomaly over the Sahel as deviations from climatology in drought years. The topic of drought-year atmospheric circulation patterns deserves more attention, so we return to it to provide suggestions for future studies in the next chapter.

Chapter 4

Conclusions and Further Study

From the onset the goals of this study have been 1) to identify a possible role of drought in global-scale vegetation changes, and 2) to verify the relationships between drought and vegetation changes on a regional scale in the Congo Basin. In progressing towards the first goal we have analyzed land type changes over the past decade and the characteristics of drought on a global scale. While statistically there is no clear correlation between variability in vegetation and tendency towards long drought, visual inspection of the results suggest that areas that have experienced land changes are also prone to long drought. This result suggests that long drought might be a mechanism for vegetation changes, although many other factors may contribute. However, the result that total months of drought does not show a pattern corresponding to regions of vegetation variability indicates that rainfall deficit alone is not a sufficient factor for inducing vegetation changes. These observations support the positive vegetation feedback system with hysteresis described in Chapter 1: isolated drought events are usually not sufficient to tip the system beyond an unstable point, but long droughts are.

In the tropics, there is evidence from station rainfall data and from satellite and model products to support the three-state vegetation stability model of Hirota et al. (2011) [17]. The observed increase in occurrence of long drought at the ends of rainfall spectra for each land type (desert, grassland, and forest) suggests that long drought is prominent in regions prone to change between land types. Thus there is an instability in between the desert, grassland, and forest land types, as seen in other studies. As discussed in the previous chapter, this result also further supports the hysteresis theory of vegetation feedback. Thus the results of this study have suggested a link between long drought and vegetation changes on a global scale. However, further work needs to be done to further validate these results.

In addressing the second goal, the results have demonstrated that forest and grassland in the Congo region interchange rather than change by other mechanisms. However, it is not clear from the studied example whether drought is linked with decreases in forest. This uncertainty is partly due to the low temporal resolution (1 year) of the land type data[1] and partly to the high variability in the LAI data. The recent drought from 2007–2012, however, is associated with a loss in forest cover. Preliminary results in atmospheric circulation for drought years in the Congo Basin have shown some features identified in previous studies with rainfall deficits. Further studies in this domain are detailed in the next section.

[1]Such a resolution is useful, however, because land can vary significantly seasonally. Additionally, the satellite measurements cannot be conducted when a region is cloudy, and so a fine resolution may be inaccurate.

This study, therefore, serves as a first step in understanding global patterns of the effects of drought on vegetation changes. There are many potential implications of such an understanding if coupled with further study in the predictability of drought and vegetation change. In particular, regions of the world that are vulnerable to vegetation change, such as areas of subsistence farming, will benefit from the ability to anticipate drought and its effects on local vegetation. The expected increase in climate variability with global warming further increases the need for such adaptive planning. Because there are so many possibilities for more research in vegetation changes, drought, and predictability, we conclude this study with several suggestions for future work to help accomplish the needs indicated above.

4.1 Further study: Atmospheric and oceanic patterns in drought years for a local study

In terms of applicability to regional drought management, studying atmospheric patterns conducive to drought is extremely important. Much like the El Niño–Southern Oscillation influences rainfall in the southeastern United States, similar patterns exist elsewhere in the world. In this study we focused on a small area of tropical central Africa. Little is known about this region, so any number of factors important in large-scale or regional atmospheric and oceanic patterns may significantly influence rainfall patterns.

A complete study of the influences on rainfall patterns in the Congo Basin or elsewhere would include many more parameters than have been analyzed in this study. Parameters such as moisture flux and divergence, sea surface temperatures, indices for oceanic temperature patterns such as the Pacific decadal oscillation (PDO) and the Atlantic multidecadal oscillation (AMO), and cloud patterns can have relevant effects on drought[2]. There are also many techniques used to determine the effects of anomalies such as the south Atlantic anticyclone on other regions of the globe. For example, analysis by empirical orthogonal functions (EOFs) can determine the propagation in time of large-scale waves from source disturbances. This work has just scratched the surface in terms of atmospheric analysis; much further study would greatly improve the understanding of drought in any specific region of interest.

4.2 Further study: Analogy of vegetation feedback to a monsoon feedback system

As mentioned in the Introduction, a recent study (Levermann et al., 2009) has explored the characteristics of the monsoon positive feedback system [26]. The moisture–advection feedback of the monsoon system exhibits many of the characterstics of instability and hysteresis that the rainfall–vegetation system does, and so it is worth studying.

In the simplest terms, a monsoon is a temperature-driven influx of moisture from the ocean to land that results in precipitation. The temper-

[2]We refer the reader to any introductory text in climate dynamics for further discussion of PDO, AMO, and other large-scale oceanic and atmospheric patterns.

ature drive arises from the unequal heating of land and water; land warms much faster during the daytime due to its lower specific heat. The temperature difference then drives winds from the ocean to the land because of the lower pressure usually associated with warm, rising air. These winds carry moisture from the ocean as well. When this humid air warms over the land, convection occurs and rainfall develops. The process of water condensation (as precipitation forms) releases latent heat over the land; this heat further increases the temperature difference between land and ocean and thus further increases the monsoon winds, allowing for more precipitation to form. Thus there is a positive feedback system involved between the incoming solar radiation, monsoon winds, and precipitation. This system is shown schematically in Figure 4.1.

Levermann et al. developed a simple mathematical model for this system such that latent heat release and heat advection balance the net radiation R, that is,

$$\mathcal{L} \cdot P - \epsilon C_p W \cdot \Delta T + R = 0, \tag{4.1}$$

where $\mathcal{L}$ is the latent heat of condensation, P is the precipitation, $\epsilon = H/L$ is the ratio between the vertical and horizontal scales of circulation, W is the horizontal monsoon wind, and ΔT is the temperature difference between land and ocean. By making several assumptions[3] and using dimensionless variables,

[3] The assumptions are that the monsoon wind W is linearly proportional to ΔT, that the precipitation balances the flux of moisture advected from the ocean, and that the rainfall P is proportional to the mean specific humidity of the atmospheric column.

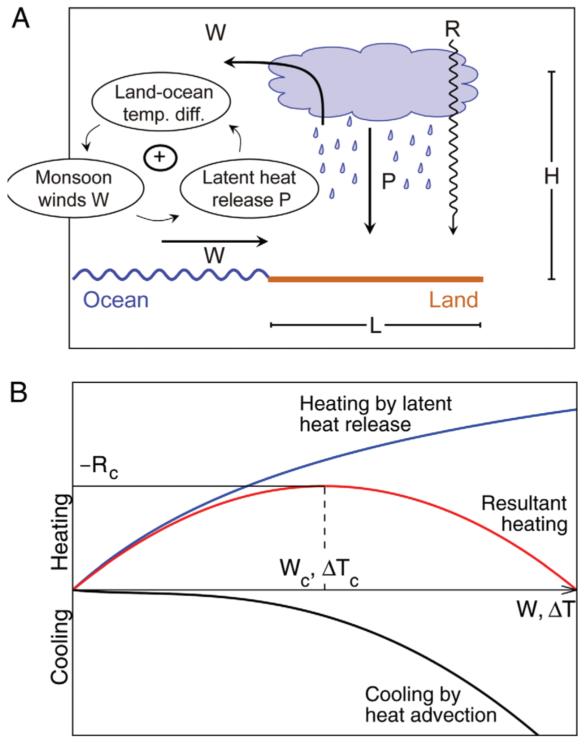

Figure 4.1: "Basic mechanism of abrupt monsoon transitions. (A) Geometry of conceptual model and fundamental moisture–advection feedback. The same notation as in the text is used for wind W, precipitation P, net radiative inux R, vertical scale H and horizontal scale L. Arrows in the feedback loop indicate the amplication of one physical processes by another. (B) Mechanism of the abrupt transition. Heating by latent heat release and cooling through heat advection compensate each other, and both decrease with decreasing winds (or equivalently, land-ocean temperature difference ΔT). The resultant heating balances the negative net radiative ux as long as it is above a threshold R_c , below which no conventional monsoon exists." Adapted from [26].

this equation becomes

$$w^3 + w^2 - (l + r)w - r = 0. \tag{4.2}$$

This equation for the dimensionless wind w depends only on the radiation r

and the parameter l, which measures the relative importance of latent and advective heat transport.

Even in a simple form this equation exhibits the properties of positive feedback, namely, a critical point beyond which the system changes abruptly, and bi-stability above the critical threshold. The simplicity and qualitative accuracy of this model suggest that a similar model may exist for the rainfall–vegetation system. The main parameters in a vegetation feedback system are the albedo of the vegetation and the effects of vegetation on the moisture balance via evapotranspiration. As noted earlier, the albedo of vegetation can increase warming by absorbing more solar radiation; this warming induces convection and thus more precipitation. Similarly, precipitation enhances the moisture of the soil in the environment and increases the amount of water that plants can retain. With higher moisture, plants can transpire more when there is no rainfall, possibly providing an influx of moisture to induce rainfall. With these mechanisms in place, it is not unreasonable to suggest that a simple mathematical model may be developed to validate the empirical observations of an unstable rainfall–vegetation feedback system.

4.3 Further study: Climate system tipping points and predictability

We conclude with a brief section concerning the predictability of abrupt changes such as those in the vegetation system. Tipping points in the climate system are defined as threshold points beyond which a forced system changes to an alternate state at a rate faster than both the internal variability of the system and the forcing rate [24]. The threshold at which the vegetation state of a region changes due to rainfall or drought is an example of such a tipping point. Several methods have been proposed as early warning mechanisms for imminent abrupt changes as a tipping point is approached [25]. These methods depend on what is known as “critical slowing down:” as a system approaches a critical tipping threshold, its ability to recover from disturbances (its resilience) is diminished. Thus by measuring how rapidly a system, such as a vegetation state, recovers from disturbances, such as minor drought, one can determine how close a system is to its tipping point.

The possibility of early warning is of great interest for many practical reasons and has many benefits in adaptation planning. However, successful implementation of an early warning system is dependent on accurate data and good knowledge of the specific system. Thus early warning is beyond the scope of this study, but will likely become an important tool in a future more extreme climate, especially for the study of drought events.

Appendix

Appendix 1

Supplemental Figures

The figures in this Appendix are referenced in the text but not considered vital to the information provided in the study. We therefore place them here with captions as sufficient explanation.

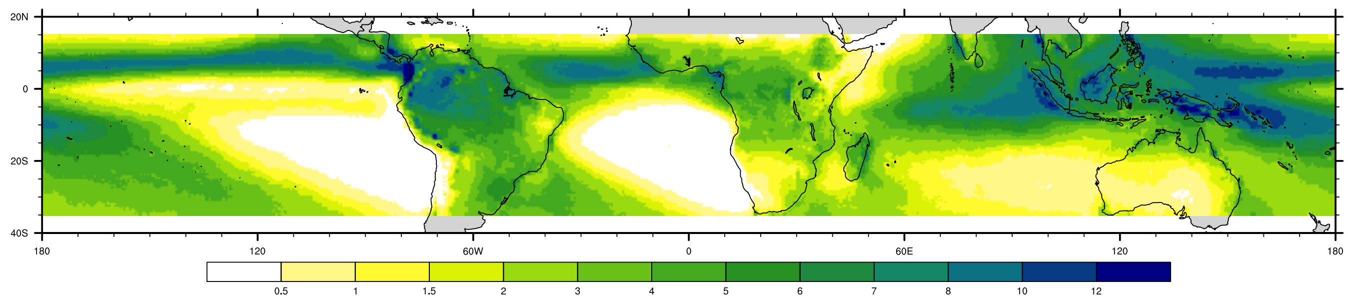

Figure 1.1: Average climatological rainfall from 1998–2012 as measured by the TRMM satellite. Units are mm day^{-1}.

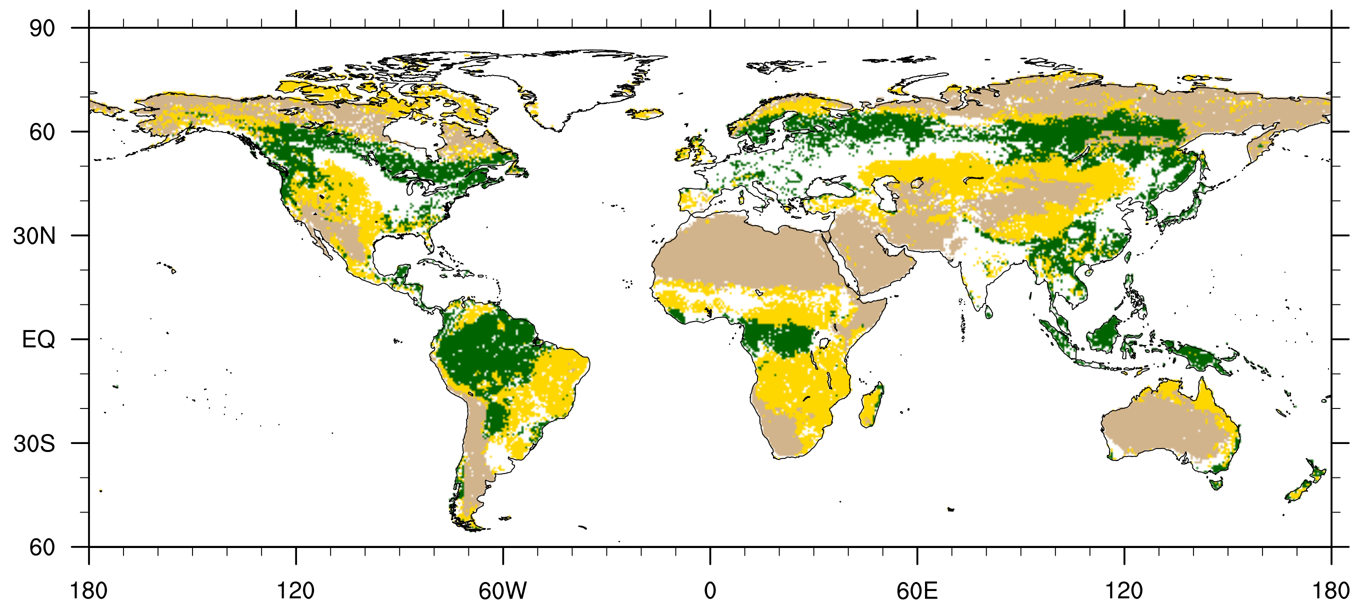

Figure 1.2: Areal coverage of dominant land type from MODIS land type data. The tan areas are predominantly desert and shrubland, the yellow are grassland, and the green are forest. Agricultural lands, wetlands, urban areas, and ice and snow are not shown.

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